Teleconnections of the Tropical Atlantic to the Tropical Indian and Pacific Oceans:

A Review of Recent Findings

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Abstract

Recent studies found that tropical Atlantic variability may affect the climate in both the tropical Pacific and Indian Ocean basins, possibly modulating the Indian summer monsoon and Pacific ENSO events. A warm tropical Atlantic Ocean forces a Gill-Matsuno-type quadrupole response with a low-level anticyclone located over India that weakens the Indian monsoon circulation, and *vice versa* for a cold tropical Atlantic Ocean. The tropical Atlantic Ocean can also induce changes in the Indian Ocean sea surface temperatures (SSTs), especially along the coast of Africa and in the western side of the Indian basin. Additionally, it can influence the tropical Pacific Ocean via an atmospheric teleconnection that is associated with the Atlantic Walker circulation. Although the Pacific El Niño does not contemporaneously correlate with the Atlantic Niño, anomalous warming or cooling of the two equatorial oceans can form an inter-basin SST gradient that induces surface zonal wind anomalies over equatorial South America and other regions in both ocean basins. The zonal wind anomalies act as a bridge linking the two ocean basins, and in turn reinforce the inter-basin SST gradient through the atmospheric Walker circulation and oceanic processes. Thus, a positive feedback seems to exist for climate variability of the tropical Pacific-Atlantic Oceans and atmospheric system, in which the inter-basin SST gradient is coupled to the overlying atmospheric wind.

1. Introduction

Sea surface temperatures (SSTs) in the tropical oceans exhibit strong interannual variations that are linked to changes in weather patterns over the continents and influence land temperatures and precipitation (e.g., Trenberth and Hurrell 1994; Latif and Barnett 1996; Thomson and Wallace 1998 for the Pacific; Vizy and Cook 2001; Giannini et al. 2003; Barreiro and Tippmann 2008 for the Atlantic; Saji et al. 1999; Webster et al. 1999; Fisher et al. 2005; Annamalai et al. 2003 for the Indian basin). The largest SST variation occurs in the tropical Pacific Ocean that hosts the El Niño-Southern Oscillation (ENSO) – arguably the most important mode of interannual climate variability on Earth. Its influence extends to all three basins and to mid- and high-latitudes by atmospheric teleconnections (e.g., Gershunov and Barnett 1998; Trenberth et al. 1998; Shukla et al. 2000; Alexander et al. 2002; Deser et al. 2004). Indeed, a correct prediction of ENSO is crucial to estimate rainfall anomalies in the Indian monsoon regions (e.g., Webster and Yang 1992), over North America (e.g., Barlow et al. 2001; Seager et al. 2005) and in the Caribbean region (e.g., Giannini et al. 2001) and over Africa (e.g., Anyamba et al. 2001). However, ENSO SSTs are not the only ones that contribute to interannual variability in the tropics, and tropical Pacific SST anomalies themselves may arise by atmospheric teleconnections from other ocean basins.

In this paper, we focus on the physical mechanisms of interannual teleconnections of the tropical Atlantic Ocean to remote regions, a topic recently investigated in a series of papers (e.g., Wang 2006; Kucharski et al. 2007; Kucharski et al. 2008a and 2008b). Notwithstanding the well established influence of the tropical Atlantic Ocean on the climate variability in its surrounding continents such as Africa and South America (e.g., Vizy and Cook 2001; Giannini et al. 2003; Janicot et al. 1998; Fontaine et al. 1998; Trazaska et al. 1996; Reason and Rouault 2006;

Barreiro and Tippmann 2008; Doyle and Barros 2002), we will show that tropical Atlantic SST anomalies can induce an atmospheric response that, in turn, modifies the oceanic and atmospheric variability in regions as far as the tropical Indian and Pacific basins. The aim is to review our recent findings on the tropical Atlantic influence on the Indo-Pacific region and to provide additional evidence for these atmospheric teleconnections.

2. Teleconnection of the Tropical Atlantic to the Tropical Indian Ocean

The influence of the Atlantic climate variability on the Indian Ocean basin has been subject of several papers, the majority of which focuses on Atlantic SSTs north of the equator. Indeed, it has been shown that changes in the Atlantic multidecadal oscillation (AMO) and in the North Atlantic oscillation (NAO) may impact the Indian monsoon (e.g. Chang et al. 2001; Lu et al. 2006; Zhang and Delworth 2006; Goswami et al. 2006; Li et al. 2008; Kucharski et al. 2008c). The physical mechanism behind the AMO-Indian monsoon teleconnection, as outlined by Goswami et al. (2006) for example, consists in an increased meridional tropospheric temperature gradient over the whole northern hemisphere including Eurasia, in correspondence to a positive phase of the AMO (i.e., positive SST anomalies in the North Atlantic). Such a temperature gradient in the troposphere may induce a late withdrawal of the Indian summer monsoon, thus causing an increase of the seasonal mean Indian monsoon rainfall. Li et al. (2008) noticed that the tropical SST expression of the AMO plays a role in this teleconnection.

Other recent studies have analyzed the interannual relation between the North Atlantic basin and the Indian monsoon (e.g., Srivastava et al. 2002; Rodwell et al. 2004; Ding and Wang 2005; Rajeevan and Sridhar 2008; Yadav 2008). Yadav (2008) and Rajeevan and Sridhar (2008), in particular, reported the existence of a correlation between south tropical Atlantic

(STA) SSTs and Indian monsoon rainfall, but they did not analyze the details of the physical mechanism. Here we discuss such a mechanism by reviewing three recent papers (Kucharski et al. 2007; Kucharski et al. 2008a and 2008b), in which Kucharski and his coauthors have shown that the influence of the Gulf of Guinea SSTs extends into the Indian Ocean basin and modulates the Indian summer monsoon on interannual timescales.

Focusing on the SST anomalies localized in the Gulf of Guinea, it is worth mentioning that the response over the African continent to their variability has been extensively discussed in the literature (Vizy and Cook 2001; Giannini et al. 2003; among others). In brief, the warming in the Gulf of Guinea forces low-level convergent flow in the surrounding coastal region, brings moist air over land, and consequently causes increased precipitation there. Kucharski et al. (2007) noted, however, that the influence of the Gulf of Guinea SSTs is not limited to the adjacent African coast, but extends further into the Indian basin and modulates the Indian summer monsoon on interannual timescales. In their work the observed interdecadal variability of the ENSO-monsoon relationship, according to which a drier than normal monsoon season usually precedes peak El Niño conditions, and *vice versa* during La Niña phase, could be realistically simulated only if changes in the tropical Atlantic SSTs in boreal summer were properly accounted for. The anti-correlation between Indian monsoon strength and the eastern equatorial Pacific SST anomalies, whose interannual variability is dominated by the ENSO signal, is well known. The physical mechanism behind it has been extensively explored in the literature (e.g., Rasmusson and Carpenter 1983; Webster and Yang 1992; Ju and Slingo 1995) and is believed to be due to the shift of the Pacific Walker circulation, with the preferred location for deep convection moving from the western Pacific warm pool to the central Pacific during El Niño events. The anti-correlation, however, is not constant over time, and undergoes decadal variability. In particular, in the late 70s, 80s and 90s this relationship has weakened substantially compared to the previous 20-25 years (Kumar et al. 1999; Torrence and Webster 1999). For example, the 1997/98 El Niño event, despite its intensity, produced only marginal rainfall anomalies over India.

Kucharski et al. (2007) investigated the observed interdecadal variability of the Indian monsoon-ENSO relationship during the period 1950-1999 using an ensemble of regionally coupled integrations. They found that an atmospheric general circulation model (AGCM) coupled to an ocean model in the Indian basin and forced by imposed SSTs elsewhere can properly simulate a dominant portion of such interdecadal variability only when observed tropical Atlantic SST anomalies are included. Before 1975 the equatorial Atlantic SSTs were slightly warmer than climatological values in correspondence to positive ENSO events and contributed to reinforce the El Niño-Indian monsoon anti-correlation. However, after 1975 a substantial cooling in the tropical Atlantic Ocean intensified the monsoonal circulation and thus weakened the ENSO-Indian monsoon relationship. As further evidenced, Kucharski et al. (2007) have shown that the simulated El Niño-Indian monsoon anti-correlation remains constant over the whole 50-year period when climatological SSTs are imposed over the tropical Atlantic (this experiment is referred to as $\mathrm{Atl}_{\mathrm{CLIM}}$). Table 1 summarizes the correlations of the June to September (JJAS) Nino3 index with an Indian monsoon rainfall (IMR) index (the IMR is defined as average JJAS rainfall over land points in the region 70°E-95°E and 10°N-30°N) and their changes before and after 1975 in the two ensembles. Note that the observed changes in the correlation coefficients are not statistically significant. Indeed, the possibility that the observed changes in the correlations are due to internal atmospheric variability has been discussed (e.g., Gershunov et al. 2001). However, the modeled correlation changes are 95% statistically significant if the individual changes in all 10-ensemble members are taken into account. As a note of caution, the reader should bear in mind that Rajeevan and Sridhar (2008) have recently observed that the correlation between the tropical Atlantic SST anomalies and Indian monsoon rainfall has weakened during recent decades, while the SST anomalies over the North Atlantic are now more strongly correlated with the IMR (see their figure 2). However, an apparent weakening of the southern tropical Atlantic (STA)-Indian monsoon relation is expected after the mid 70's if the ENSO and STA influences are not separated in the analysis, due to the competing effect of tropical Pacific and Atlantic SST anomalies on the Indian monsoon. In other words, positive SST anomalies in both basins act to decrease the Indian monsoon, but if the SSTs in the two basins are anti-correlated, as in recent decades, the net impact of tropical SSTs outside the Indian basin on the Indian monsoon will be reduced because the contributions from the Atlantic and Pacific tend to cancel each other.

The Atlantic SST anomalies, significantly correlated with the Indian monsoon rainfall, have been found to extend between 30°W to 10°E and 20°S to 0°S using both observational data and coupled models (Kucharski et al. 2008a), and are usually referred to as southern tropical Atlantic (STA) pattern (Huang et al. 2004). This pattern is very similar to the SST regression onto the so-called Atl3 index, defined as averaged SST anomalies in the region of 20°W to 0°E and 4°S to 4°N (e.g., Zebiak 1993). The correlation coefficient of the Atl3 and STA SST anomalies is indeed 0.89.

The physical mechanism by which the tropical Atlantic influences the Indian monsoon has been investigated by Kucharski et al. (2008b) with idealized AGCM experiments. In these simulations, SST anomalies are prescribed only in the Gulf of Guinea region and climatological SSTs force the AGCM elsewhere (see Fig. 1). The magnitude of the SST anomaly shown in Fig.

1 is twice that of a regression of SSTs onto the STA index. 100-ensemble members of seasonal integrations for the season of July to September (JAS) have been performed. The atmospheric response to the equatorial heating anomaly is of a Gill-Matsuno-type (Gill 1980; Matsuno 1966). In the summer season, there are indeed upper-level easterly wind anomalies near the equator in the Atlantic and Indian Ocean region, which favor a pure equatorially trapped Gill-Matsuno response, consistent with the findings of Jin and Hoskins (1995). Rossby waves are responsible for transporting the signal to the west, while Kelvin waves carry the westerly wind anomalies to the east. The large-scale response to the SST anomaly of Fig. 1 in 200-hPa streamfunction is a quadrupole with the northeastern upper level trough being located just west of the Indian Peninsula (Fig. 2a). This quadrupole has a baroclinic structure with a low-level anticyclone located over India, causing low-level divergent motion and a reduction of precipitation over the Indian region (Fig. 2b).

The equatorial wind response is determined by a low-level convergence in the tropical Atlantic and West African region that leads to upward motion and consequently upper-level divergence. This, in turn, causes upper-level westerly wind anomalies to the east and easterly wind anomalies to the west. Being a baroclinic response, at low levels there are equatorial easterly wind anomalies in the Indian Ocean region (Fig. 2b).

Here we also present some new evidence for the Atlantic-Indian monsoon interaction, focusing on the impact of STA SST anomalies on winds and SSTs in the Indian basin. We begin our new analysis by considering the JJAS dynamical Indian monsoon index (IMI), defined as the north-south 850-hPa zonal wind difference averaged over the domains of 40°E-80°E, 5°N-15°N and 60°E-90°E, 20°N-30°N and derived from the NCEP/NCAR reanalysis (Kalnay et al. 1996). The IMI is representative of the Indian monsoon strength (e.g., Wang et al. 2001). A similar analysis has been also presented in Kucharski et al. (2008b), but for the JAS season. Figure 3a shows the correlation coefficients (CCs) of the IMI with the SSTs from the HadISST dataset (Rayner et al. 2003) for the period 1950 to 1999. CCs larger than 0.3 are 95% statistically significant. As already mentioned, the existence of a strong anti-correlation between the Indian monsoon strength and the eastern equatorial Pacific SST anomalies associated with ENSO is well established (Fig. 3a). The correlations seen in other regions, on the other hand, should be interpreted with care, as ENSO-induced variability may force at the same time both the IMI and the remote SST anomalies, and correlations may be present without a physical connection between IMI and those SSTs. Therefore, using a technique similar to that adopted in Kucharski et al. (2008a), we define the residual IMI, IMI_{RES}, as the IMI component that is not linearly related to ENSO according to:

$$IMI_{RES}(t) = IMI(t) - IMI_{ENSO}(t)\,, \tag{1}$$

$$IMI_{ENSO}(t) = bNINO34(t)\,, \tag{2}$$

where $NINO34(t)$ is the Nino3.4 index defined as area averaged SST anomalies in the region of 170°W-120°W and 5°S-5°N and b is calculated with a least square linear regression. Figure 3b shows the CCs of IMI_{RES} with SSTs for the period 1950 to 1999. Clearly, Gulf of Guinea SST anomalies display the largest correlations, while all other regions are characterized by hardly statistically significant CCs. This analysis supports the previous findings on the influence of tropical Atlantic SSTs over the Indian basin using wind measurements.

Finally, we notice that the tropical Atlantic–Indian basin teleconnection induces changes in Indian Ocean SSTs as well. As previously discussed, a warm anomaly in the STA region of the tropical Atlantic induces low-level easterly wind anomalies in the western tropical Indian Ocean (Fig. 2b). These wind anomalies, in turn, affect the Indian Ocean SSTs. We investigated this response in both reanalysis datasets and in regionally coupled model integrations. We use, once more, the ensemble described in Bracco et al. (2007) and Kucharski et al. (2007) and realized with a global AGCM coupled to an ocean model in the Indian basin, while prescribed observed SSTs force the atmospheric model elsewhere. Figures 4a and b show the regressions of SSTs and 850-hPa winds onto the JJAS STA index for observations (HadISST for SST and NCEP/NCAR re-analysis for winds) and model data, respectively. Note that the result is very similar if the ENSO signal is removed from the STA index because the correlations of the STA and NINO3 indices are very small for the period considered here (CC = 0.08 for 1950 to 1999). Consistent with a weakening of the Somali jet, SSTs in the western tropical Indian Ocean are warmer than climatological values due to decreased evaporation and ocean downwelling in response to warm STA SST anomalies (with values of about 0.2°K per standard deviation of the STA index) both in the observations (Fig. 4a) and in the coupled model ensemble (Fig. 4b). The corresponding correlation coefficients exceed 0.4 in the western Indian Ocean, thus explaining an important portion of the Indian Ocean SST variability on interannual timescales. The link of the response of the Indian Ocean to the STA SST anomalies with the Indian Ocean zonal mode variability in autumn (e.g., Saij et al. 1999) needs to be further investigated.

3. Teleconnection of the Tropical Atlantic to the Tropical Pacific

Tropical Atlantic variability is dominated by two climate modes: The Atlantic zonal equatorial mode and the tropical Atlantic meridional gradient variability (e.g., Xie and Carton 2004; Chang et al. 2006). The Atlantic zonal equatorial mode is also called the Atlantic Niño.

Here we discuss and summarize the influence of the tropical Atlantic on the tropical Pacific through the inter-basin SST gradient that is formed by the equatorial modes of the Pacific El Niño and the Atlantic Niño (Wang 2006). The Pacific El Niño and the Atlantic Niño are measured by the Nino3 (150°W-90°W, 5°S-5°N) and Atl3 (20°W-0°, 4°S-4°N) SST anomalies, respectively (e.g., Neelin et al. 1998; Wang and Picaut 2004; Zebiak 1993). As pointed out by many studies (e.g., Zebiak 1993; Enfield and Mayer 1997), the Nino3 SST anomalies do not contemporaneously correlate with the Atl3 SST anomalies. However, the inter-basin SST gradient does not require a contemporaneous correlation between the Nino3 and Atl3 SST anomalies. The inter-basin SST gradient can be formed when anomalous events of two oceans are either of the same sign or opposite sign as well as in one ocean basin alone. This is somehow analogous to the tropical Atlantic meridional SST gradient variability that is associated with the tropical North and South Atlantic SST anomalies which are mostly independent, however.

The inter-basin SST gradient index is defined by the difference between the standardized Nino3 and Atl3 SST anomalies (Wang 2006), representing what is happening in both the equatorial eastern Pacific and Atlantic Oceans. Figure 5 shows the correlation maps of 850-hPa wind anomalies, SST anomalies and sea level pressure (SLP) anomalies with the inter-basin SST gradient index. When the SST anomalies are positively correlated in one ocean, the correlation in the other ocean basin is negative (Fig. 5b). Correspondingly, the SLP gradient pattern is established with a negative correlation in the eastern tropical Pacific and a positive value in the tropical Atlantic (Fig. 5c). Since zonal SLP gradient determines surface zonal wind near the equatorial region (Lindzen and Nigam 1987), the east-to-west (or westward) SLP gradient results in equatorial easterly wind anomalies over equatorial South America (ESA) and the equatorial Atlantic (Figs. 5a and 5c). In the Pacific, the west-to-east (or eastward) SLP gradient is associated with strong equatorial westerly wind anomalies (Figs. 5a and 5c), consistent with the wind distribution of Pacific warming. The inter-basin SST gradient can induce the overlying atmospheric wind anomalies across ESA, which can help bridge the interaction of the two ocean basins.

The cross-South America zonal wind anomalies can in turn induce the changes in the equatorial eastern Pacific and Atlantic Oceans. The oceanic responses of sea surface height (SSH) and SST anomalies to the ESA surface zonal wind anomalies are shown in Fig. 6. SSH variation is a proxy for variability of the thermocline, with a positive (negative) SSH anomaly implying a deepened (shallowed) thermocline. Figure 6a indicates that equatorial easterly (westerly) wind anomalies over ESA are associated with the deepened (shallowed) thermocline in the equatorial eastern/central Pacific and the shallowed (deepened) thermocline in the equatorial eastern/central Atlantic. The variations of the thermocline are consistent with the fact that the ESA easterly (westerly) wind anomalies correspond to warm (cold) SST anomalies in the equatorial eastern Pacific and cold (warm) SST anomalies in the equatorial Atlantic (Fig. 6b).

Since the equatorial eastern Pacific Ocean and the equatorial Atlantic Ocean are separated by the landmass of northern South America, their interaction must be through induced variability of the atmosphere. Figure 7 shows the correlation maps of the upper tropospheric (200-hPa) velocity potential and divergent wind anomalies and rainfall anomalies with the inter-basin SST gradient index. These variations indicate that the zonal Walker circulations driven by the zonal heating asymmetry are responsible for the coherent picture of the equatorial Pacific and Atlantic Oceans. Associated with the inter-basin SST variability are two zonal anomalous Walker circulation cells: The Pacific and Atlantic Walker cells (Fig. 8). The air anomalously ascends in the equatorial central/eastern Pacific and flows both westward and eastward in the upper troposphere. The westward flow of the upper troposphere is associated with the Pacific Walker circulation that induces the Bjerknes feedback for ENSO events. The eastward flow of the upper troposphere corresponds to the Atlantic Walker circulation that is responsible for the inter-basin interaction. Over the equatorial Atlantic and western Pacific, the air anomalously converges and descends. At the lower troposphere, the air then anomalously converges toward the equatorial central/eastern Pacific and forms the two Walker circulation cells (Klein et al. 1999; Wang 2002a, 2002b and 2005).

The observational results suggest a positive feedback mechanism in the tropical Pacific and Atlantic in which the cross-South America zonal wind variability is coupled to the SST gradient between the equatorial eastern Pacific and Atlantic (Fig. 9). Suppose that there is a positive SST anomaly in the equatorial eastern Pacific or a negative SST anomaly in the equatorial Atlantic. The initial zonal SST gradient between the two equatorial ocean basins induces an east-to-west pressure gradient in the atmospheric boundary layer (Lindzen and Nigam 1987), which results in equatorial easterly wind anomalies across northern South America. The cross-South America easterly wind anomalies extend eastward to the equatorial Atlantic and cool the equatorial Atlantic Ocean through oceanic dynamics (e.g., Zebiak 1993; Carton and Huang 1994) and then further enhance the initial inter-basin SST gradient. In the Pacific side, the cross-South America easterly wind anomalies (extending to the west coast of northern South America as shown in Fig. 5a) may increase lower (upper) tropospheric convergence (divergence) in the equatorial eastern/central Pacific, and thus enhance the anomalous Walker circulation in the Pacific. The strengthening of the anomalous Pacific Walker circulation, added to that of the Bjerknes positive feedback within the Pacific, is associated with an increase of SST in the equatorial eastern/central Pacific. This also seems to enhance the initial inter-basin SST gradient. It thus suggests the existence of a feedback loop that involves the interaction between the inter-basin SST gradient and overlying atmospheric flow across northern South America.

We have to point out that this data-inferred possible feedback needs to be confirmed by using coupled ocean-atmosphere models. However, as suggested by one reviewer, here we examine how the atmosphere responds to SST anomalies of one ocean basin. Figure 10 shows the AGCM responses of rainfall, surface wind, surface pressure and 200-hPa velocity potential to a cold SST anomaly in the tropical Atlantic similar to the pattern shown in Fig. 1. The model results are consistent with the above observational data and the proposed hypothesis. The cold SST anomalies in the equatorial Atlantic produce negative (positive) rainfall anomalies in the equatorial Atlantic (eastern Pacific) and easterly wind anomalies across ESA (Fig. 10a). Both surface pressure and 200-hPa velocity potential show positive values in the tropical Atlantic and negative values in the tropical eastern Pacific (Figs. 10b and c). All of these are consistent with that an inter-basin SST gradient can produce an overlying atmospheric flow which can bridge the two ocean basins.

4. Summary

Among three of the tropical oceans, the vast tropical Pacific Ocean hosts the largest interannual phenomenon of ENSO whose influence spans the globe. Not surprisingly, most of studies have been focused on the teleconnections of Pacific ENSO to the tropical Atlantic and Indian Oceans during the past decades. Recently, some studies have emphasized the potential teleconnections of the tropical Atlantic to the tropical Indian and Pacific Oceans (e.g., Wang 2006; Kucharski et al. 2007; Kucharski et al. 2008a and 2008b). Both observational data and numerical model show that the tropical Atlantic Ocean can teleconnect to the Indian basin influencing precipitation, winds, SLP and SSTs. A warming (cooling) in the Gulf of Guinea region forces a Gill-Matsuno-type quadrupole response with a low-level anticyclone (cyclone) located over India that weakens (strengthens) the Indian monsoon circulation. Because of this relationship, the breakdown of the ENSO-Indian monsoon relation observed in the last decades may be attributed to the interdecadal variability of tropical Atlantic SSTs. Additional and new analyses in section 2 further show that a warm (cold) tropical Atlantic Ocean can induce a warm (cold) SSTs in the Indian Ocean, especially along the coast of Africa and in the western side of the Indian basin.

Observations support the notion that the tropical Atlantic can influence the tropical Pacific Ocean via the inter-basin SST gradient variability that is associated with the Atlantic Walker circulation. Both the tropical Pacific and Atlantic Oceans host an equatorial mode of interannual variability called the Pacific El Niño and the Atlantic Niño, respectively. Although the Pacific El Niño does not contemporaneously correlate with the Atlantic Niño, anomalous warming or cooling of the two equatorial oceans can form an inter-basin SST gradient variability that induces surface zonal wind anomalies over equatorial South America and over some regions of both ocean basins. The zonal wind anomalies act to bridge the interaction of the two ocean basins, reinforcing the inter-basin SST gradient through atmospheric Walker circulations and oceanic dynamics. Thus, a positive feedback seems to exist for climate variability of the tropical Pacific-Atlantic Oceans and atmosphere system, in which the inter-basin SST gradient is coupled to the overlying atmospheric wind. Rainfall responds to the inter-basin SST gradient by showing an anti-symmetric configuration between the two equatorial oceans, suggesting that rainfall is sensitive to the equatorial inter-basin SST gradient, regardless of which ocean is anomalously warm or cold. An AGCM experiment, forced by the tropical Atlantic SST anomalies, is consistent with the data-inferred mechanisms described here. However, the proposed feedback involved the interaction between the equatorial Pacific and Atlantic Oceans needs to be further investigated by using coupled ocean-atmosphere models.

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Table 1: Correlation Coefficients (CC) between IMR and the Nino3 index for CRU (the first row), ensemble mean of Atl_{ANOM} (the second row), and ensemble mean of Atl_{CLIM} (the third row) for the periods 1950 to 1999, 1950 to 1974, and 1975 to 1999. The last column shows the CC difference of the later minus the earlier period for CRU and the respective experiments.

	1950-1999	1950-1974	1975-1999	1975-1999 minus 1950-1974
CC (CRU, Nino3)	-0.59	-0.69	-0.45	0.24
CC (Atl_{ANOM}, Nino3)	-0.63	-0.79	-0.51	0.28
CC (Atl_{CLIM}, Nino3)	-0.73	-0.74	-0.72	0.013

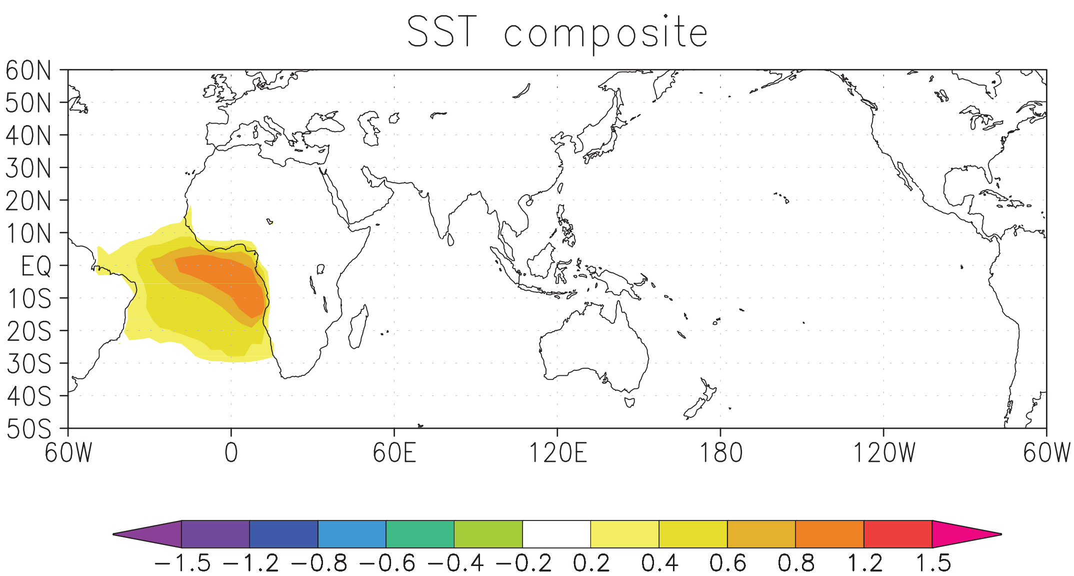

Figure 1. Composite SST anomaly. See text for details. Units are °K.

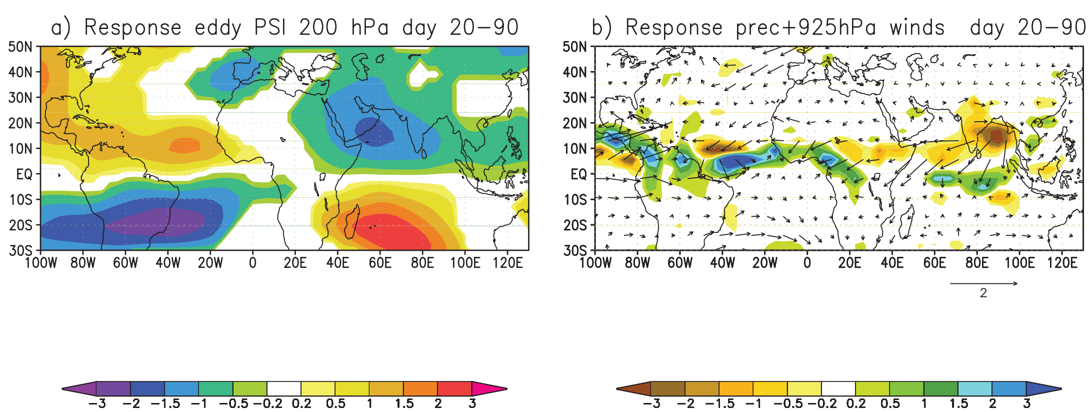

Figure 2. Response of (a) 200-hPa eddy streamfunction and (b) precipitation and 925-hPa winds to a south tropical Atlantic SST anomaly (see Fig. 1). All shown contour intervals are 95% statistically significant. Units are 10^6 m^2/s for (a) and mm/day for precipitation and m/s for wind in (b).

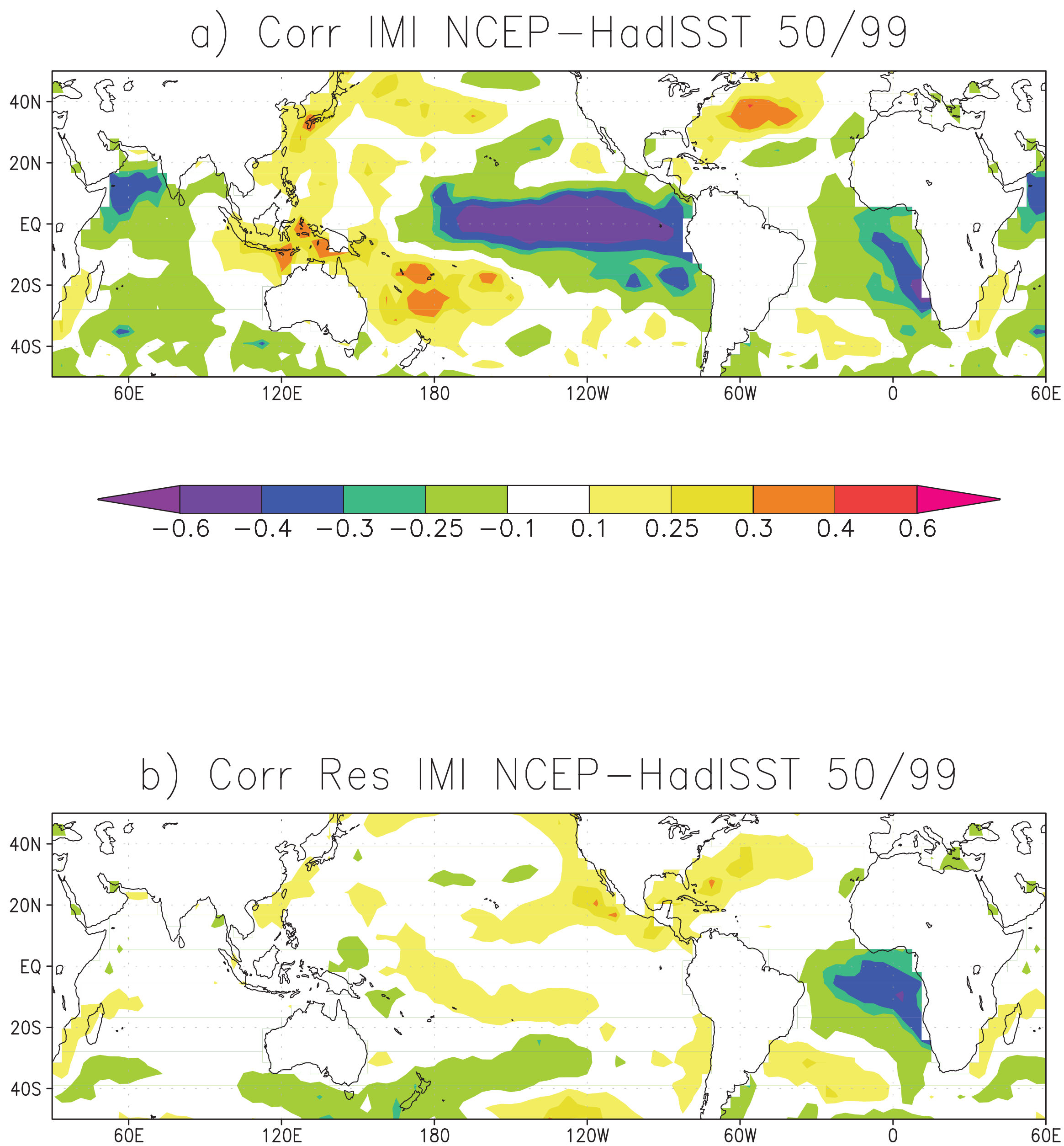

Figure 3. Distribution of the Correlation Coefficient (CC) between (a) IMI and observed SSTs and (b) IMI_{RES} and observed SSTs. Absolute CC lagrer than 0.3 are 95% statistically significant.

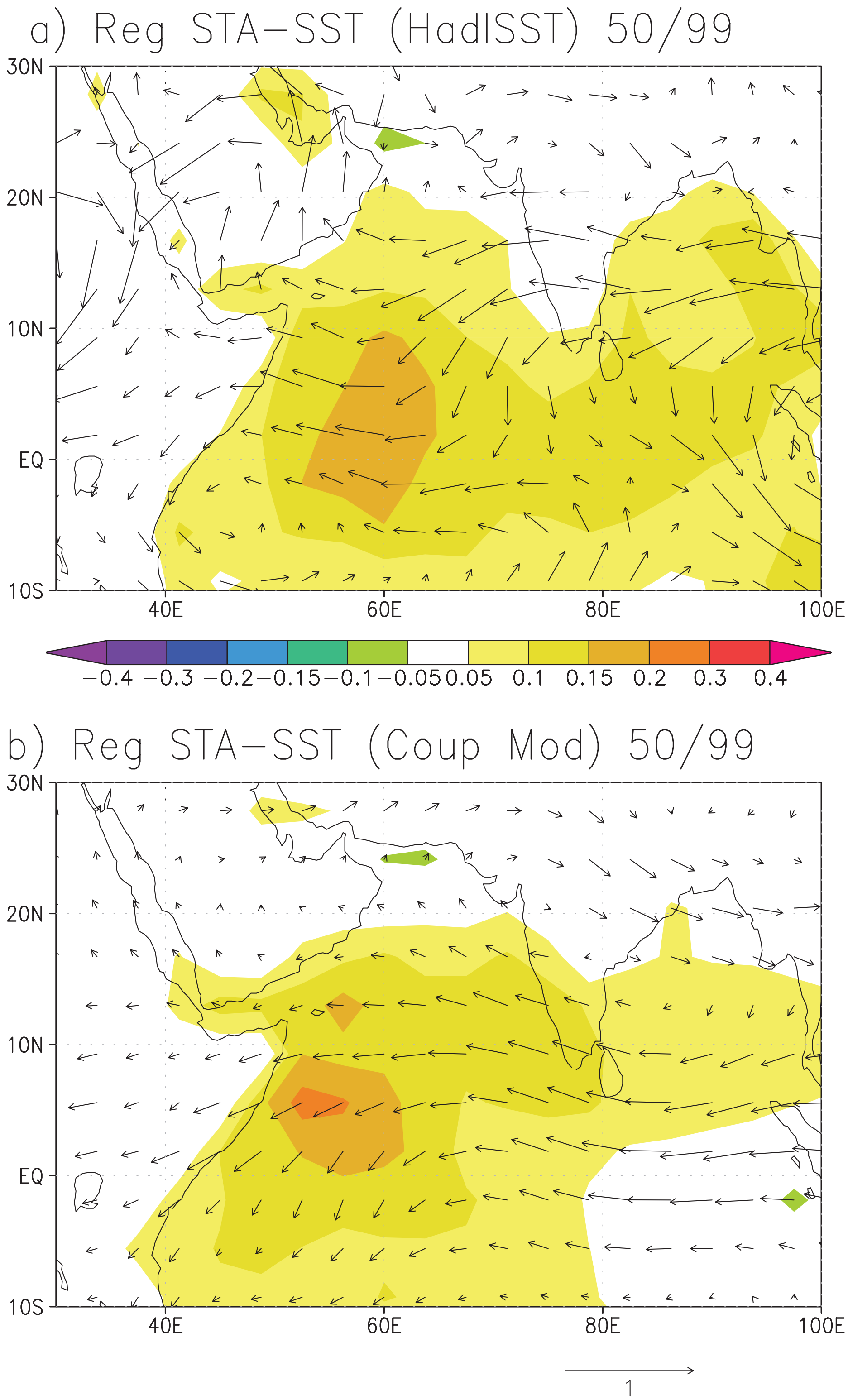

Figure 4. Regression of (a) observed and modeled (b) SSTs and 850-hPa winds onto the STA index. Units are °K and m/s per STD of the STA index.

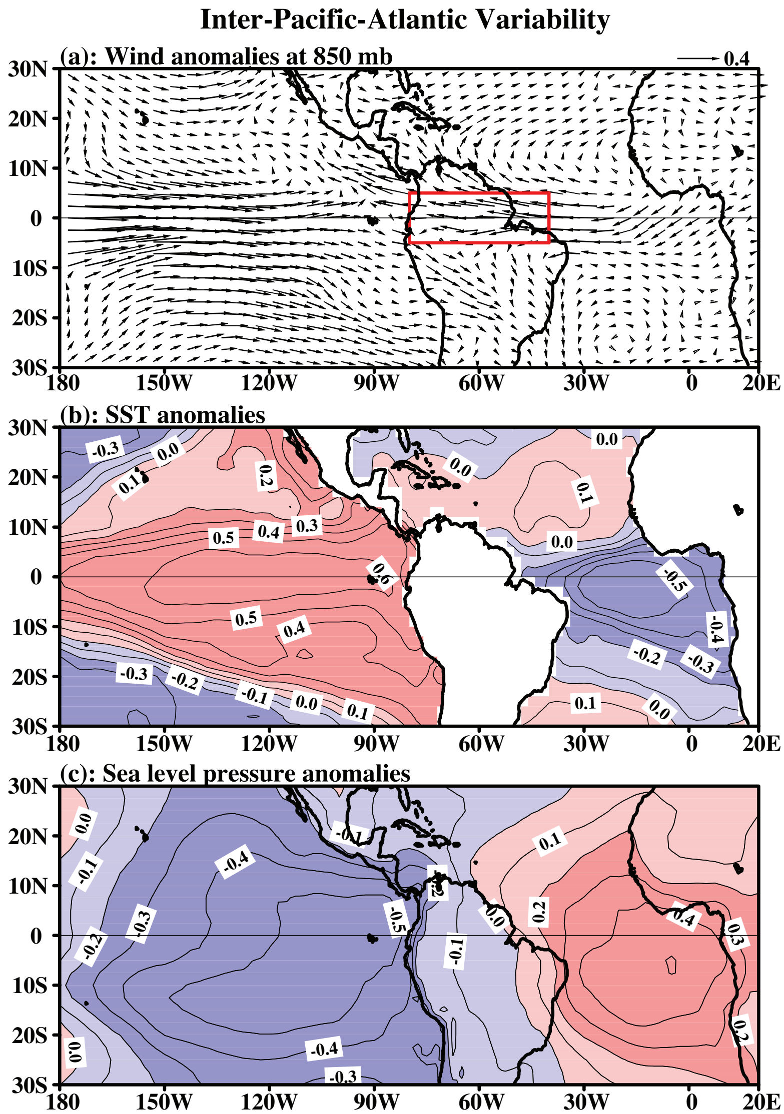

Figure 5. Correlation maps of (a) wind anomalies at 850-hPa, (b) SST anomalies, and (c) sea level pressure anomalies with the inter-basin SST anomaly gradient index. The red rectangular box in (a) represents the region of equatorial South America (ESA). In (b) and (c), positive (negative) correlation is in red (blue) color. The dark red and dark blue represent correlation larger than 0.2 (95% significant level). All data are monthly and from 1950 to 2004.

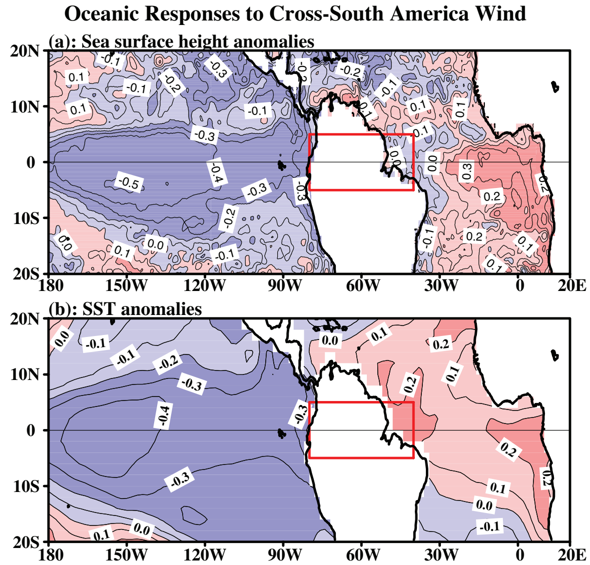

Figure 6. Oceanic responses to the cross-South America zonal wind anomalies. Shown are correlation maps of (a) sea surface height (SSH) anomalies and (b) SST anomalies with the 850-hPa zonal wind anomalies in the equatorial South America region of 5°S-5°N, 80°W-40°W (red rectangular box). Positive (negative) correlation is in red (blue) color. The dark red and dark blue represent correlation larger than 0.2 (95% significant level). Monthly SSH is from 1993-2004 and other data are from 1950 to 2004.

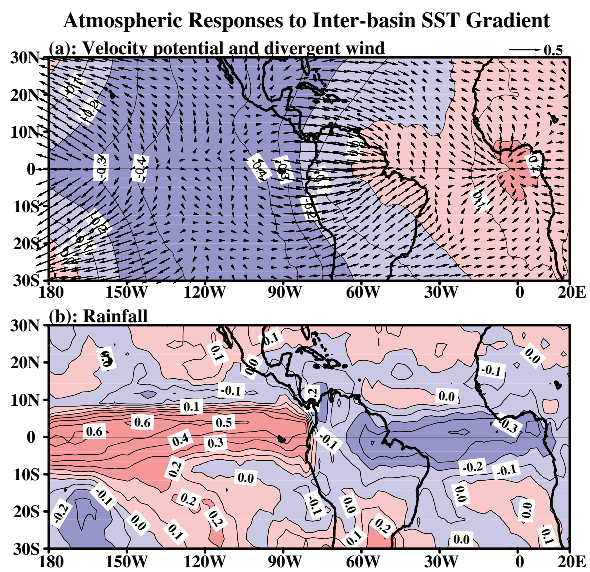

Figure 7. Atmospheric responses to the inter-basin SST gradient variability. Shown are correlation maps of (a) the 200-hPa velocity potential and divergent wind anomalies and (b) rainfall anomalies with the inter-Pacific-Atlantic SST anomaly gradient index. Positive (negative) correlation is in red (blue) color. The dark red and dark blue represent correlation larger than 0.2 (95% significant level). Monthly rainfall is from 1979-2004 and other data are from 1950 to 2004.

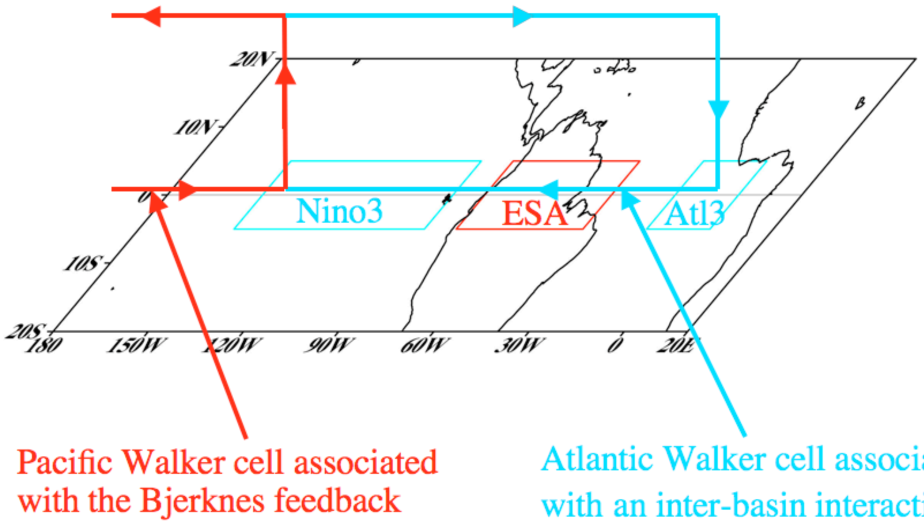

Figure 8. Schematic diagram showing two anomalous Walker circulations: The Pacific Walker cell and the Atlantic Walker cell.

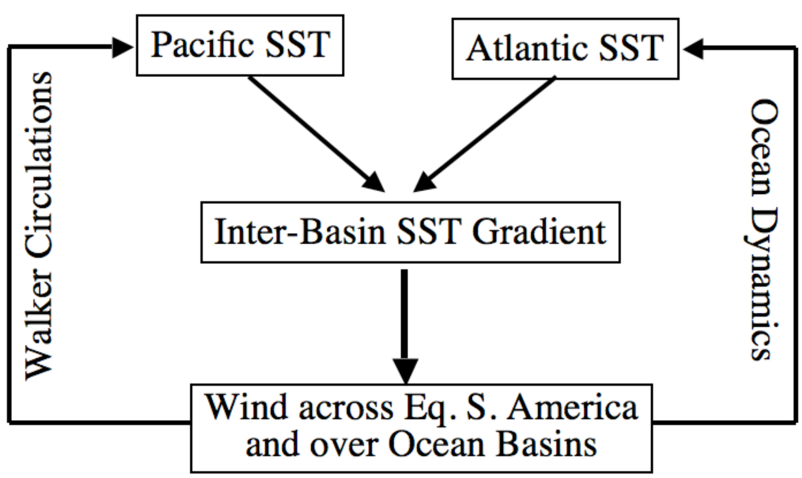

Figure 9. Schematic diagram showing that a positive feedback loop involving the interaction between the inter-basin SST gradient and overlying atmospheric flow across northern South America.

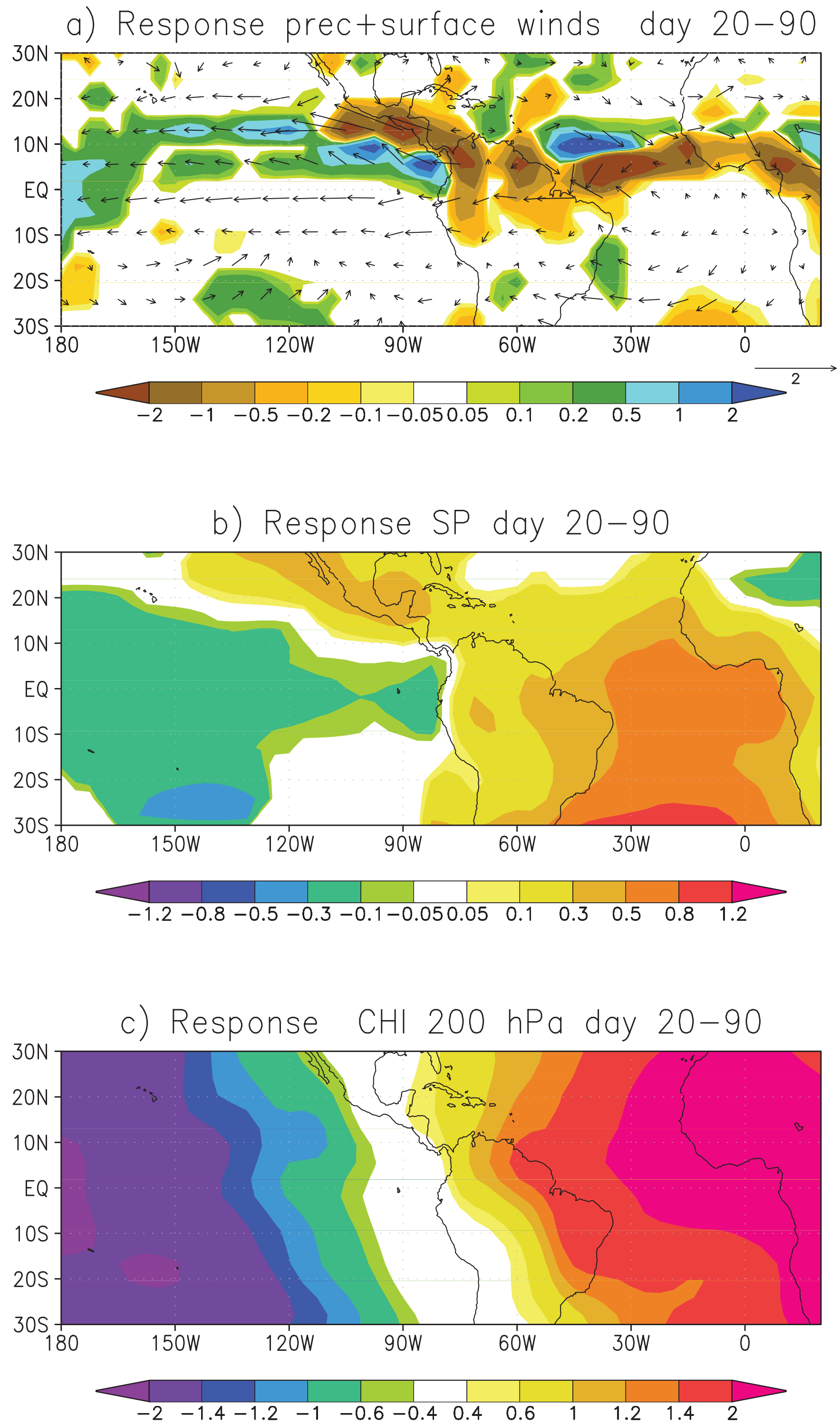

Figure 10. AGCM responses of (a) rainfall (mm/day) and surface wind (m/s), (b) surface pressure and (c) 200-hPa velocity potential to a cold SST anomaly in the south tropical Atlantic that is similar to the pattern shown in Fig. 1.